

Forearc collapse, plate flexure, and seismicity
within the downgoing plate along the Sunda
Arc west of Sumatra
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Abstract

Deformation within the downgoing oceanic lithosphere seawards of subduction zones is typically characterised by regimes of shallow extension and deeper compression, due to the bending of the oceanic plate as it dips into the subduction zone. However, offshore Sumatra there are shallow compressional earthquakes within the downgoing oceanic plate outboard of the region of high slip in the 2004 Aceh-Andaman earthquake, occurring at the same depth as extensional faulting further seaward from the trench. A clear separation is seen in the location of intraplate earthquakes, with extensional earthquakes occurring further seawards than compressional earthquakes at the same depth within the plate. The adjacent section of the forearc prism west of Aceh is also anomalous in its morphology, characterised by a wide prism with a steep bathymetric front and broad, gradually-sloping top. This shape is in contrast to the narrower and more smoothly-sloping prism to the south, and along other subduction

zones. The anomalous near-trench intraplate earthquakes and prism morphology are likely to be the result of the geologically-rapid gravitational collapse of the forearc, which leads to induced bending within the subducting plate, and the distinctive plateau-like morphology of the forearc. Such collapse of the forearc could be caused by changes through time of the material properties of the forearc rocks, or of the thickness of the sediments entering the subduction zone.

Highlights:

- Near-trench intraplate compressional seismicity is observed in the downgoing plate
- Earthquakes are indicative of near-trench unbending of the lithosphere
- Seismicity and forearc morphology are consistent with gravitational forearc collapse

Keywords: Sumatra, intraplate seismicity, forearc deformation, flexure

1 Introduction

On 24th December 2004, the M_W 9.2 Aceh-Andaman earthquake ruptured a section of the subduction interface along the Sunda arc stretching from Simeulue island, west of Sumatra, northwards to the Andaman islands, ~ 1300 km along strike (Figure 1; Ammon et al., 2005; Rhie et al., 2007; Chlieh et al., 2007). Most major subduction-interface earthquakes are followed by the widespread rupture of normal faults in the downgoing plate seawards of the trench (e.g., Lay et al., 1989, 2009; Craig et al., 2014a). These earthquakes are the result of the release of shallow extensional stresses in the outer rise region of the downgoing plate as it bends into the subduction zone. However, the 2004 Aceh-Andaman earthquake is so far unique in the observational record in that it was followed by shallow compressional, rather than extensional, seismicity beneath the trench and under the outer trench slope/outer rise, along with only a small number of normal-faulting aftershocks within

47 the downgoing plate (Dewey et al., 2007).

48 The near-trench compressional seismicity offshore Sumatra has variously
49 been interpreted as the transfer of the active subduction interface from the
50 top of the downgoing plate into the mantle of the downgoing plate (Singh
51 et al., 2008), as a shallow response within the downgoing plate to high levels
52 of induced stress at the updip termination of the 2004 mainshock rupture
53 on the interface, or as shallow motion on splay faults branching up from
54 the main interface (Dewey et al., 2007). However, correctly understanding
55 the tectonic significance of these earthquakes relies on accurately estimat-
56 ing their locations, depths, and mechanisms. The determination of accurate
57 estimates for the location of these intraplate earthquakes, at a resolution be-
58 yond routine global seismological techniques, is therefore of vital importance.
59 Similarly, one of the most accurate ways of constraining the location of the
60 active subduction megathrust – critical for determining which earthquakes
61 are truly intraplate – is through the precise location of low-angle thrust-
62 faulting earthquakes that lie on this interface. In the first part of this study,
63 we therefore present the results of body-waveform modelling to constrain the
64 source parameters of the near-trench seismicity offshore Sumatra (Figure 1),
65 in order to image the deformation field within the downgoing oceanic plate.

66 In the second part of this study, we investigate the links between our
67 seismological results and the structure and morphology of the forearc prism.
68 The Sunda Arc is notable for both its variable forearc morphology along
69 strike (McNeill and Henstock, 2014), and major along-strike variations in
70 the thickness of sediments on the downgoing plate (Figure 1e, see also com-
71 piled data in Table 1 of McNeill and Henstock 2014). Incoming sediment
72 thickness varies from 1 – 5 km, with the greatest thickness occurring along a
73 section of the trench stretching north from Simeulue island (2.6°N, 96.0°E)
74 to approximately 6.5°N, and overlaps with the region of highest slip in the
75 2004 earthquake. In this area, west of northern Sumatra, the forearc is char-
76 acterised by a wide forearc prism with a relatively low-gradient top and steep
77 frontal slope (Figure 2e – h), in contrast to the region to the south (Figure
78 2i,j) where the prism is characterised by the more gently-sloping rise from
79 the trench over a wider across-strike extent.

80 The morphology and internal structure of a forearc prism is controlled
 81 by a number of competing factors, including the dip and physical properties
 82 of the subduction interface, the material properties of the over-riding acce-
 83 tionary wedge, the thickness and character of incoming sediments, and the
 84 degree to which they are accreted onto the frontal prism, underplated onto
 85 the base of the prism, or subducted along with the downgoing plate. Whilst
 86 the growth and evolution of accretionary prisms is often treated as being
 87 uniform through time, we investigate how changing some of the properties
 88 governing its shape (specifically, incoming sediment thickness or internal rhe-
 89 ology) can lead to a relatively rapid readjustment in the prism shape, which
 90 also leads to a concurrent adjustment of the induced stress field within the
 91 downgoing plate. We then present a conceptual model linking the morpholog-
 92 ical evolution of the forearc prism to the changing stress distribution within
 93 the downgoing plate, as mechanism to explain both the anomalous prism
 94 morphology and the unique distribution of seismicity.

95 2 Seismicity

96 2.1 Modelling

97 We here determined earthquake source parameters for events along the Sunda
 98 arc, in close proximity to the trench, by the inversion of long-period body
 99 waves using the algorithm of Zwick et al. (1994). The workflow followed is
 100 similar to that described in detail in Tilmann et al. (2010) and Craig et al.
 101 (2014a). Teleseismic P - and SH -waves were inverted over a time window
 102 encompassing the direct arrival (P , S) and subsequent principal depth phases
 103 (pP , sP , sS) to determine the source mechanisms, centroid depth and seismic
 104 moment of earthquakes with $M_W \geq 5.5$ since 1990. Examples are shown in
 105 Supplementary Figures 1 and 2.

106 For events occurring seawards of the trench, a source-side velocity struc-
 107 ture was used consisting of a crustal layer 7 km thick ($V_P = 6.5 \text{ ms}^{-1}$, $V_S =$
 108 3.8 ms^{-1} , and $\rho = 2800 \text{ kg m}^{-3}$) over a mantle halfspace ($V_P = 8.1 \text{ m s}^{-1}$,
 109 $V_S = 4.6 \text{ m s}^{-1}$, and $\rho = 3300 \text{ kg m}^{-3}$). To compensate for the laterally

110 varying thickness and seismic velocity structure of the accretionary wedge
 111 landward of the trench, the crustal layer thickness is increased and the ve-
 112 locities and density reduced with increasing distance from the trench, in-line
 113 with the results of refraction profiles across the region (Dessa et al., 2009;
 114 Singh et al., 2012). In each case, this velocity structure is overlain by a wa-
 115 ter layer with the water depth in the source region being initially based on
 116 the SRTM30PLUS bathymetric models (filtered to remove wavelengths of less
 117 than 10 km), and adjusted if required to best fit any observed water multiples,
 118 although the inversion window is limited where possible to exclude further
 119 water multiples after the sP arrival. The inclusion of horizontally-polarised
 120 S -waves aids in minimising the effects of any inaccuracies (or azimuthal vari-
 121 ability) in water depth on the depth determination of the earthquake, as the
 122 horizontal polarisation excludes and converted P -wave phases from featuring
 123 in the waveform coda. The restricted frequency content of the long period
 124 P -wave data also reduces the P wave sensitivity to water depth (Engdahl
 125 and Billington, 1986).

126 Direct P - and S -wave arrivals were manually picked from broadband seis-
 127 mograms in each case. The earthquakes modelled are shown in Figure 1 and
 128 listed in Table S1, and include the majority of events with $M_W \geq 5.5$ occur-
 129 ring in the study area within 400 km of the trench. The exceptions are in the
 130 period immediately following the mainshock ruptures of the Aceh-Andaman
 131 and Nias earthquakes, in late December 2004 and late March 2005 respec-
 132 tively, as the signals from smaller-magnitude aftershocks during the initial
 133 hours after the mainshocks were swamped by the mainshock coda, and failed
 134 to yield robust results.

135 Typical uncertainties in source mechanism are on the order of 10° for
 136 strike and rake, and 5° for dip (e.g. Molnar and Lyon-Caen, 1989; Taymaz
 137 et al., 1991; Craig et al., 2014b). Depth uncertainties, of most direct relevance
 138 to this study, are usually $\sim \pm 3$ km (Tilman et al., 2010), much of which
 139 derives from the velocity model used. Hence, relative uncertainties between
 140 earthquakes in the same geographic location are often smaller. Accounting
 141 for increased uncertainty in the depth estimates due to bathymetric variation
 142 around the source, and the differing effect this has on the depth phases

143 for stations with different bouncepoints, we estimate a further increase in
144 uncertainty for events near sharp bathymetric variations of ~ 1 km for the
145 deepest of our studied events, although we note that due to the increasing
146 moveout of the depth-phase bouncepoints with increasing source depth, this
147 uncertainty is itself depth-dependent.

148 Whilst the focus of this work is on deformation in the downgoing plate, it
149 is also necessary to determine source parameters for low-angle thrust-faulting
150 aftershocks associated with motion on the main subduction interface, so as to
151 correctly define the location of this interface, and to determine whether events
152 were in the downgoing plate or within the overlying accretionary wedge.
153 Hence, a large number of the low-angle thrust-faulting earthquakes shown
154 on Figures 1a & 1d are in fact on the plate interface, and not within the
155 downgoing plate. To supplement these events in determining the location of
156 the plate interface, we also draw upon a detailed study of large-magnitude
157 interface aftershocks at the southern end of the study area that was conducted
158 by Tilmann et al. (2010), along with three microseismic surveys conducted
159 in the aftermath of the major interface events of 2004 and 2005 (black points
160 on Figures 2a and 3; Lin et al. 2009; Lange et al. 2010; Tilmann et al. 2010).
161 In using the results from local seismic networks, we only show earthquakes
162 located within the area covered by the network, and well constrained events
163 that are based on observations at multiple (≥ 5) stations of both P and S
164 arrivals.

165 2.2 Earthquake distribution

166 Seismic activity in the study area is shown on Figure 1, and is dominated
167 by thrust-faulting earthquakes, many of which show a low-angle, northeast-
168 dipping nodal plane consistent with motion on the main subduction interface
169 around the margins of the mainshock slip patch (Figure 1e). Mechanisms in
170 the area around the boundary between the 2004 and 2005 source regions,
171 previously determined by Tilmann et al. (2010), are all also consistent with
172 low-angle thrust-faulting seismicity on the subduction interface (indicated by
173 the larger green points on Figure 1a).

174 A large number of low-angle thrust-faulting earthquakes also occur be-
 175 neath the Aceh basin region (Figure 1d). Previous studies have suggested
 176 that these may represent motion on a splay fault (Waldhauser et al., 2012) or
 177 the reloading and repeat rupturing of small asperities within a section of the
 178 interface otherwise undergoing aseismic afterslip (Yu et al., 2013). However,
 179 whilst we do find a slight deepening of these earthquakes with distance from
 180 the trench, we find insufficient difference between the depths and mechanisms
 181 of these earthquakes to distinguish between these possible causes.

182 There are also a number of thrust-faulting mechanisms beneath and sea-
 183 wards of the trench with orientations (in particular, dip angles) that are
 184 inconsistent with motion on a low-angle subduction interface (Figure 1a).
 185 Whilst these earthquakes are found in a range of locations along the trench,
 186 a major concentration occurs at $\sim 2.5^\circ\text{N}$, with a range of focal mechanism
 187 orientations, and depths of 6 – 26 km below the seafloor (Figure 1a). This
 188 cluster lies to the south of the region of highest slip in the 2004 mainshock
 189 (Figure 1e), and in the region of thickest sediment on the incoming plate
 190 (Figure 1e), and is the main subject of the next section.

191 In contrast to the widespread thrust-faulting earthquakes, normal-faulting
 192 mechanisms are sparse (Figure 1b), with only 10 near-trench normal-faulting
 193 events with $M_W > 5.5$, nine of which have occurred since the 2004 mainshock
 194 are present in our catalogue, and all of which are indicative of bending-driven
 195 horizontal extension within the shallow outer-rise or outer-trench slope re-
 196 gion as observed in other subduction zones (Christensen and Ruff, 1988;
 197 Craig et al., 2014a). Three normal-faulting earthquakes have also occurred
 198 significantly landward of the trench, one indicating deeper extension within
 199 the downgoing plate, and two indicating extension at the base of the forearc,
 200 at depths within error of the inferred plate interface.

201 In the last decade, there have been a number of major strike-slip earth-
 202 quakes located in the interior of the Indian plate, including the M_W 8.7
 203 April 2012 earthquake (e.g. Yue et al., 2012), associated with a region of
 204 diffuse deformation in the Wharton basin (Delescluse et al., 2012; Aderhold
 205 and Abercrombie, 2016). Strike-slip seismicity in our study region, both sea-
 206 wards and landwards of the trench, follows a general trend of NNE-SSW and

207 ESE-WNW aligned nodal planes (Figure 1.c). The alignment of the approx-
 208 imately north-south nodal planes with oceanic fracture zones in this region
 209 (Figure 1.c,f), and the identification of lineations in microseismic activity
 210 beneath the accretionary wedge (Lange et al., 2010), indicate the widespread
 211 reactivation of the pre-existing oceanic fabric, both seaward and landward
 212 of the trench, consistent with a detailed study of Indian Ocean seismicity in
 213 this region (Aderhold and Abercrombie, 2016). Two of the events on Fig-
 214 ure 1.c, at $\sim 5.75^\circ$ N 93.25° E, lie along-strike from the 2012 Indian Ocean
 215 earthquakes, and may represent continued deformation of the same fracture
 216 zone beneath the accretionary wedge. Shallower strike-slip seismicity land-
 217 wards of the trench is concentrated along the Sumatran and West Andaman
 218 fault systems, which accommodate the strike-slip component of the oblique
 219 convergence between the Indian Ocean and Sunda.

220 Little conclusive evidence is seen within the region for large-scale seis-
 221 mic activity within the forearc prism, outside of these major strike-slip sys-
 222 tems (Figure 1 and Figure 3). Previous studies have suggested that the
 223 accretionary wedge in this region may undergo either gravity-driven exten-
 224 sion and collapse (McKenzie and Jackson, 2012), or compressional motion
 225 on splay faults following the mainshock (Chauhan et al., 2009). Both of
 226 these mechanisms might be expected to be expressed in the seismicity within
 227 the accretionary prism. Of the normal-faulting earthquakes analysed here,
 228 none locate conclusively within the accretionary prism, although the depths
 229 of only a small number of the normal-faulting earthquakes recorded in the
 230 gCMT catalogue (blue triangles, Figure 1b) for the forearc region could be
 231 confirmed using the waveform modelling techniques employed here. This
 232 difficulty arises because many of these events occurred in the time period
 233 directly following the mainshock, when continuing seismic coda from the
 234 mainshock prevents a robust inversion using bodywaves. A single high-angle
 235 thrust at 3.9° N, 95.3° E is confirmed to occur at a depth placing it in the
 236 accretionary prism (see Figure 1a), and this might represent seismogenesis
 237 on recently active splay faults within the prism (Graindorge et al., 2008;
 238 Chauhan et al., 2009), but it is unclear how widespread such deformation
 239 is. Presently-available seismological observations are therefore not able to

unambiguously constrain the orientation of the principal strains within the accretionary wedge. However, the small number of earthquakes imply that much of the strain is likely to be accommodated aseismically.

2.3 Seismicity within the downgoing plate

As described above, much of the seismicity offshore Sumatra represents motion on the subduction interface. However, the near-trench intraplate seismicity, particularly the cluster of thrust-faulting earthquakes at $\sim 2.5^\circ\text{N}$, present an important contrast to globally observed patterns of seismicity within the downgoing plate of subduction zones (Chapple and Forsyth, 1979; Christensen and Ruff, 1988; Craig et al., 2014a). On a global scale, normal faulting seaward of the trench is typically observed from the surface of the downgoing plate down to some transition depth, below which the plate either becomes aseismic, or switches to thrust-faulting earthquakes. This pattern, with shallow extension overlying deeper compression, is consistent with the accumulation of horizontal extensional strain along the top of the strong lithospheric plate, and horizontal compression along the base, as the plate itself bends into the subduction zone. Such bending-related strain, although accommodated by seismogenic brittle failure on faults, is expected to be recovered further on in the subduction process, as the subducting slab returns to being roughly planar as it descends into the upper mantle. This unbending of the slab downdip of the interface seismogenic zone is a common interpretation of the focal mechanisms of double seismic zones downdip of the seismogenic subduction interface (Engdahl and Scholz, 1977; Kao and Chen, 1996; Gamage et al., 2009). The location of the transition between bending and unbending is difficult to constrain, but in most subduction zones where it can be observed, it occurs significantly landward of the trench, and shallow normal-faulting earthquakes indicative of horizontal extension due to bending persist from the outer rise region to the trench (Craig et al., 2014a).

On Figures 2a and 3, we separate the seismicity of the subducting system into three geographic sections (divided by the green dotted lines on Figure 2a), and plot earthquake depth profiles as a function of distance to the trench

for each section (Figure 3). We analyse the seismicity of each of these profiles from south to north in turn, and assess how they compare with the global pattern of seismicity within the downgoing plates at subduction zones:

1. South of 1.5°N (Figure 3c) the plate interface (approximated by the grey lines on Figure 3) is clearly delineated by a line of low-angle thrust-faulting earthquakes (see Figure 1a and Tilmann et al. 2010). The two normal-faulting events in this area are both trench-parallel. The shallowest one is consistent with bending-related extension seaward of the trench. The deeper event, located just landward of the trench, may indicate either that extension extends to 30 km into the plate, or may indicate a transition to unbending (with extension at the base of the plate), as the plate straightens out under the forearc. It is interesting to note that the location of both extensional earthquakes (in depth and across-strike distance) is matched by a cluster of microearthquakes imaged by Lange et al. (2010).
2. Between 1.5°N and 6.5°N (Figure 3b), a more complex pattern of seismicity is seen. The subduction interface is clearly delineated at > 50 km from the trench by a combination of low-angle thrust-faults (both those beneath the Aceh basin, and others further south) and microseismic aftershocks beneath Simeulue (green circles; Tilmann et al., 2010). At ~ 50 km seaward from the trench, a single shallow normal-faulting earthquake at 6 km depth is consistent with the typical model for shallow extension due to outer-rise bending (Chapple and Forsyth, 1979; Christensen and Ruff, 1988). Beneath this, a thrust-faulting earthquake at 40 km is consistent with compression in the deeper part of the bending plate, but the orientation of this mechanism is near-perpendicular to the trench, possibly instead reflecting along-strike curvature of the plate as the trench changes strike west of Northern Sumatra. In close proximity to the trench itself, seismicity is characterised by widespread thrust-faulting, extending from the surface down to > 30 km. This observations is, to our knowledge, unique in the world's subduction zones during the instrumental period (Craig et al.,

2014a). The widespread depth extent of these earthquakes is inconsistent with the idea that they might be concentrated onto a single low-angle structure (the subduction interface), and some of them must represent brittle failure in horizontal compression within the upper sections of the downgoing plate. The juxtaposition of these thrust and normal earthquakes shows a horizontal transition from shallow extension (the normal fault) to shallow compression (the thrust faults) as the trench is approached, as discussed below.

3. North of 6.5°N (Figure 3a), more sparse thrust-faulting earthquakes again serve to illuminate the subduction interface. In the near-trench region, two clusters of normal-faulting earthquakes at shallow depth (< 25 km) within the oceanic plate occur with mechanisms sub-parallel to the trench, indicating bending-related faulting. Thrust-faulting earthquakes within 20 km of the trench occur at depths of 10 – 20 km, and with steeper dips than the interface events further landward. The depth extent over which we find these thrust-faulting earthquakes, and the variability in the orientation of their mechanisms (see Figure 1a) is inconsistent with all of them being focused on the main plate interface. However, the true interpretation of these events is uncertain – their depths suggest deformation similar to that seen over a larger depth interval on Figure 3b, and suggest that at least some of these earthquakes lie in the upper part of the downgoing plate. However, the more limited depth extent, and the lack of thrust-faulting earthquakes deeper than 16 km, means that we cannot rule out the possibility that these earthquakes represent either near-trench splay faulting, or compression in the frontal section of the forearc accretionary prism.

In summary, the southern section of our study area shows seismicity consistent with the globally-observed pattern for outer-rise regions, of bending-related shallow extension. The area west of Aceh, however, does not, and is instead characterised by the occurrence of thrust-faulting earthquakes within 20 km of the trench (both landwards and seawards) at a range of depths from 6 km to over 30 km (Figure 3b). This observation is inconsistent with inter-

pretations that these earthquakes all occurred on the subduction interface, that they occurred on shallow splay faults branching upwards from the interface, or that they represent internal deformation within the toe of the forearc prism. The northernmost section of our study area may fit with the trend seen west of Aceh, but more limited seismicity, along with moderate-depth extension in the downgoing plate, mean we cannot rule out other deformation scenarios.

The shallow compressional seismicity within the downgoing plate occurs in an area where bathymetric surveys show evidence for well-developed trench-parallel normal faults breaking the top surface of the downgoing plate (Cook et al., 2014). The near-juxtaposition of these contrasting deformation indicators suggests a change in deformation through time, from the extension that produced the bathymetric scarps, to the presently-active faults that can be seen in the earthquake activity. Whilst the stress state within the downgoing plate is expected to vary, up to a point, across the interface seismic cycle, no evidence has been found elsewhere in the world for an outer rise region failing in both extension and compression either side of a major earthquake on the adjacent interface, despite an exhaustive search of recent outer-rise seismicity (Craig et al., 2014a). Additionally, the vast majority of the seismicity included in our study occurs in the years following the 2004 and 2005 interface events (see Supplementary Figure 2), at a time in the interface seismic cycle when the stress state within the downgoing plate oceanwards of the interface rupture patch is expected to be at its most extensional. The temporal evolution of stress is therefore presumably a longer-term effect, beyond the timescales of individual megathrust earthquake cycles.

This apparently-flexural seismicity within the downgoing plate is distinct from the intraplate deformation seen within the Wharton Basin (Wiens et al., 1985; Delescluse and Chamot-Rooke, 2007; Carton et al., 2014). This is particularly clear when considering the difference between the orientation of P- and T-axes for the near-trench thrust faulting, and the strike-slip faulting that dominates the internal deformation of the Wharton Basin. P-axes for the strike-slip faulting are orientated roughly NNW-SSE – approximately parallel to the strike of the subduction zone. In contrast, P-axes for the

368 near-trench thrust-faulting earthquakes are orientated ENE-WSW, roughly
 369 perpendicular to that seen in the strike-slip faulting. We hence consider the
 370 causative process behind the near-trench seismicity to be distinct from that
 371 leading to the diffuse intraplate deformation of the Wharton basin. 2D seis-
 372 mic reflection studies have indicated the presence of small-offset faults within
 373 the Indian Ocean plate SW of Aceh (Carton et al., 2014), likely penetrating
 374 down into the oceanic mantle. Given the limitations of 2D seismic survey-
 375 ing, the orientation and true dip of these faults remains uncertain. However,
 376 their location and probable moderate dip angle suggests that they are not
 377 compatible with the deeper thrust-faulting seismicity discussed here, which
 378 occurs at either steep or shallow dip angles (depending on which nodal plane
 379 is the true fault plane), and closer to the trench.

380 **3 Forearc evolution and stresses in the down-** 381 **going plate**

382 The highly unusual oceanic intraplate seismicity described above occurs in
 383 a location also noted for its unusual forearc morphology, discussed in detail
 384 elsewhere (Kopp et al., 2008; McNeill and Henstock, 2014; Moeremans et al.,
 385 2014; Cook et al., 2014). Figure 2 shows across-strike averaged bathymetric
 386 profiles through a range of trench-perpendicular swaths, shown on Figure 2a,
 387 consistent with the available prism transects of ship-board bathymetry and
 388 2D seismic data (see Figure 4 of McNeill and Henstock 2014). In the region
 389 of shallow oceanic intraplate compression ($1.5^{\circ}\text{N} - 6.5^{\circ}\text{N}$), the forearc shows
 390 a distinctive and unusual shape with a relatively flat top and sharp, steeply-
 391 sloping wedge-front (see Figure 2) characterised by the presence of landward-
 392 vergent folds (Henstock et al., 2008; McNeill and Henstock, 2014; Cook et al.,
 393 2014). In comparison, to the south of this region, the forearc shows the more
 394 commonly-observed shape of a relatively smoothly-sloping prism front from
 395 the trench up onto the prism top (Figure 2b-d). Additionally, following the
 396 definitions of McNeill and Henstock (2014), wherein the prism is defined
 397 as extending from the trench to edge of the forearc basin (often bounded

by a margin-parallel fault system) the total prism width in this region is significantly wider (~ 150 km) than is to the north or south (~ 100 km). The relatively flat plateau top through this region typically comprises 100 – 140 km of this total width. This leads to a prism with a distinct, sharp change in gradient $\lesssim 50$ km landwards from the trench. In contrast, the section to the south is characterized by a much narrower prism ($\lesssim 120$ km), with a gently curved slope profile (Figure 2i,j).

The northern section (Figure 2b–d) shows an extremely wide prism with a low angle, gradually sloping prism front. Given the ambiguous nature of both the seismicity and prism morphology of this northern section, likely complicated by the increasing proximity to both the Andaman spreading centre and the Bengal fan, we do not discuss it further here, but instead focus on the difference between the central and southern sections, and the transition between them near 1.5°N .

Next, we describe a dynamic model which is designed to investigate the potential causes of the unusual intraplate seismicity and forearc morphology. Based on the prevalence of ductile deformation features within the forearc wedge (i.e. folds), and the absence of significant seismicity, we model the forearc wedge using a viscous rheology (which is what would result from fluid-assisted pressure-solution/diffusion creep in the thick sedimentary pile (e.g. Rutter, 1983)). We will initially describe some simple two-parameter models that capture the governing physics of the accretionary wedges, before discussing a more complex multi-parameter thermomechanically-coupled model of our suggested mechanism for the evolution of the Sumatra forearc.

In our models, the accretionary wedge is underlain by the subduction megathrust, which we model as a constant-shear-stress lower boundary to the deformation within the wedge. The model consists of convergence between the rigid oceanic plate, and a deformable sedimentary veneer, with a rigid ‘backstop’ that represents the rigid part of the over-riding plate, against which the internally-deforming forearc prism builds a forearc wedge from the accumulation of the incoming deformable sediment (the model geometry is shown in Figure 4a). We solved the equations for low-Reynolds number fluid flow using the finite-difference methods described in Reynolds et al. (2015).

431 We non-dimensionalise the equations for Stokes flow using the thickness
 432 of sediment on the downgoing plate as the length-scale (H on Figure 4a),
 433 and the incoming plate velocity (u_0 on Figure 4a). The deformation is then
 434 governed by the equations

$$\nabla' h' = \alpha \nabla'^2 \mathbf{u}' \quad (1)$$

$$\alpha = \frac{\eta u_0}{\rho g H^2} \quad (2)$$

436 where h is the surface elevation, \mathbf{u} is the velocity vector, η is the prism
 437 viscosity, ρ its density, g the gravitational acceleration, and primes denote
 438 non-dimensional quantities. In our model, we then solve of Eq. 1 in cross
 439 section only. α is analogous to the inverse of the Argand number (commonly
 440 used to described the viscous deformation of continental collision zones; Eng-
 441 land and McKenzie 1982), and represents the ratio of the stresses required
 442 to deform the wedge and the gravitational forces acting upon it. The other
 443 quantity in our model setup is the shear stress on the base of the wedge (τ_m ,
 444 non-dimensionalised as $\tau'_m = \tau_m H / \eta u_0$), which appears as the lower bound-
 445 ary condition on our model domain. Where the shear stress on the bottom
 446 boundary is below τ_m , the sediments remain mechanically attached to the
 447 downgoing plate (i.e. a horizontally-rigid lower boundary condition), and de-
 448 form by internal shearing of the sedimentary package. Where the shear stress
 449 reaches τ_m , the boundary condition is imposed such that there is sliding on
 450 the fault at the base of the wedge, with the velocity required for the shear
 451 stress on the base of the overlying material to equal τ_m .

452 The growth of the forearc wedge is a balance between the stresses on the
 453 base (τ_m) that are able to support the overlying topography, and gravity
 454 acting to reduce the elevation of the wedge by lateral spreading. If τ and
 455 α remain constant through time, the balance between these effects leads
 456 to a wedge that grows in a close to self-similar manner. This situation is
 457 the viscous equivalent of a ‘critical taper’ coulomb wedge. Such a model
 458 is shown in Figure 4b for the case where the stresses on the subduction
 459 thrust dominate the growth of the prism, with little deformation occurring
 460 in response to topographic forces until when the prism height is roughly five
 461 times larger than the incoming sediment thickness (upper line on the figure).

462 In our modelling approach we can investigate the lateral and temporal
463 variations in the style of strain that result from changes in the model param-
464 eters. Figure 4c shows the effect of reducing the value of α by a factor of 10,
465 with the starting topography in the model given by the red line in Figure 4b.
466 The wedge undergoes gravitational collapse, the front rapidly advances, and
467 the topography develops a low-gradient top and a steeper front. The rate
468 of propagation of the prism slows down as a new dynamic balance between
469 the forces acting upon it is approached. It is therefore clear that changes in
470 the value of α can result in rapid transient propagation of the wedge, and a
471 change in the overall morphology.

472 A number of effects could change the value of α (Eq. 2). The most
473 likely reason for a dramatic change in α is due to a change in the viscosity of
474 the wedge. In shallow sedimentary sections, the viscosity for rocks deform-
475 ing by solution-precipitation creep (i.e. diffusion creep) is highly dependent
476 on temperature, and so on depth. This effect arises because the viscosity
477 is governed by an Arrhenius relation, as with other creep mechanisms (i.e.
478 $\eta = A \exp(-E/RT)$, where A is a constant, E is the activation energy, R is
479 the gas constant, and T is temperature) (Rutter, 1983; Connolly and Pod-
480 ladchikov, 2000). In slowly-deposited deep-sea sediments, the thermal profile
481 is in equilibrium, so depth is a proxy for temperature. The exponential term
482 in the expression for viscosity can lead to dramatic changes in viscosity over
483 small depth intervals. For example, Connolly and Podladchikov (2000) mod-
484 elled a decrease in viscosity of over 1.5 orders of magnitude between depths of
485 1 and 2 km. The appearance of dramatically lower-viscosity sediments being
486 input into the wedge, because of kilometre-scale increases in the incoming
487 sedimentary thickness, would make dramatic changes to the average viscos-
488 ity of the wedge on short timescales, and could lead to the effects modelled
489 above because of the dramatic reduction in α .

490 Decreases in the rate of convergence with time could also reduce the
491 value of α . This effect would reduce the rate of sediment input, and so lead
492 to collapse of the wedge. However, it is unlikely that the convergence in
493 Sumatra has changed significantly in recent times (DeMets et al., 2010), and
494 such a change would affect the entire arc, rather than only one section of it.

There are unlikely to be major temporal changes in the density of the wedge because of the limited variation in the density in the incoming sediments, which is considerably less and an order of magnitude. The thickness of the incoming sediments appears in the expression for α , as a separate effect from the thermal and viscosity effects discussed above. A sudden change in α could be interpreted as a change in the incoming sediment thickness. However, because H enters into the expression for α as $1/H^2$, and the viscosity depends on $\exp(-E/RT)$, where $T \propto H$, we are likely to be in a regime where the exponential term is more dominant than the quadratic, and so the viscosity effects discussed above are more important in this setting.

Changing the value of τ_m (basal shear stress) can also lead to the outwards growth of the prism. However, this occurs as a shallowing of the roughly constant-gradient wedge front seen in 4b, failing to produce a steep front to the evolving prism (Figure S4), and therefore is less consistent with the morphology of the Sumatra forearc west of Aceh than decreasing the value of α .

The gravitational collapse of the wedge as shown in Figure 4 will affect the stress-state of the underlying oceanic plate (Figure 5). If the outwards propagation of the wedge is more rapid than the rate at which the subducted slab can ‘roll back’ through the mantle, the wedge collapse and the propagation of the collision front out over the incoming plate will result in the zone of bending moving ocean-wards, and the creation of a region of opposite-polarity un-bending close to the nose of the wedge. In this location, where the oceanic plate flattens under the propagating thrust belt, previously accrued extensional strain is recovered through shallow compression within the downgoing plate (Figure 5). Changing α therefore provides a mechanism to explain both the highly unusual oceanic intraplate seismicity and the distinctive forearc morphology offshore Sumatra. The precise nature of the induced stress field remains uncertain, due to the rheological complexity of the downgoing plate, and remaining uncertainties in the response of faults to applied stresses. However, given the magnitude of the change in the overriding topography, the stresses produced are likely to be on the order of 100’s MPa – far greater than observed stress drops in intraplate earthquakes, and there-

fore easily sufficient to influence the pattern of bending-related deformation and seismicity that we observe.

The simple two-parameter model discussed above captures the dominant controls on the behaviour of accretionary wedges, without the added parameters that arise in a fully thermomechanically-coupled model. In order to demonstrate this point, in the supplemental information we include a model for the evolution of the temperature and deformation within the forearc in which thermomechanical coupling has been implemented (see Figure S5). The complexity of this model, in terms of the wide range of free parameters with unknown values, means that it does not provide any additional insights into the evolution of accretionary wedges. However, it is included to demonstrate that the results of our two-parameter model, which point towards collapse of the Sumatran forearc in response to an influx of thick, hot, and weak sediment, are mirrored by more complex models.

4 Controls on forearc equilibrium

The question remains as to which of the potential controlling factors (prism viscosity or incoming sediment thickness) may have changed significantly in the geologically recent past in the region of Aceh. Internal prism viscosity is expected to evolve over time as the prism builds up, changing its internal thermobarometric state. However, this evolution will proceed slowly, on the timescale of prism formation, and the prism geometry would be expected to evolve gradually to maintain an equilibrium with the evolving viscosity (see Figure S5). The presence of anomalous intraplate seismicity in the outer rise region, along with the development of the unusual forearc morphology, suggests a more rapid gravitationally-driven collapse.

The input of relatively warm and low-viscosity sediments into the wedge, due to a change in sediment thickness on the incoming plate, provides a mechanism for the prism to undergo rapid collapse. Incoming oceanic sediment thickness is largely a function of three parameters: plate age (and hence pelagic sediment thickness), proximity to clastic sediment sources, and geographic relation to basin-bounding features (e.g., fracture zones). In the case

559 of the Sunda Arc, variation in clastic sediment input and composition are
 560 relatively small along strike, south of the region of influence of the Bengal
 561 fan, which reaches down to the approximate latitude of the Nicobar islands
 562 ($\sim 11^\circ\text{N}$). Although the age of the incoming plate varies across our study
 563 area by approximately 30 Myrs, the dominant influence on sediment thick-
 564 ness is the structural segmentation of the downgoing plate by fracture zones,
 565 and the major features of the Ninety East ridge and the fossil spreading ridge
 566 that intersects the trench at $\sim 0.5^\circ\text{N}$. Figure 1e summarises the known con-
 567 straints on the sediment thickness at the trench (McNeill and Henstock, 2014,
 568 and references therein), and demonstrates that sediment thickness along this
 569 section of the arc varies from as low as 1 – 2 km at the northern and southern
 570 ends of our study area, to as high as 4 – 5 km in the central section west of
 571 Aceh, also characterised by the anomalous forearc morphology, and shallow
 572 compression within the downgoing plate.

573 **5 Gravitational signature of prism collapse**

574 The gravitationally-driven collapse of the forearc prism should be evident
 575 in gravity data, and indeed marine free-air gravity anomalies in the region
 576 also suggest that this region of the forearc is anomalous (Figure 1f). Gravity
 577 profiles across the trench typically show a wide gravity low centred on the
 578 trench itself, associated with the flexure of the downgoing plate, followed by
 579 a gradual rise to a gravity high at the peak of prism, as seen in the profiles for
 580 central Sumatra shown in Figure 2i,j. West of Aceh, however, the negative
 581 gravity anomaly associated with incoming plate flexure decays rapidly, and
 582 the profile rises sharply in the region of the trench itself, reaching a relative
 583 high ~ 40 km landward of the trench (Central Section, Figure 2). The
 584 gravity profile then returns to a strong negative anomaly further landwards,
 585 over the low-gradient section of the forearc prism. The near-trench positive
 586 anomaly and prism-top negative anomaly match the gravity field expected
 587 for a region undergoing collapse due to gravitationally driven instability, as
 588 mass is rapidly moved from the wedge top to wedge front at a rate faster
 589 than the underlying plate can re-adjust.

590 In keeping with the uncertain nature of the near-trench seismicity in the
 591 northern section of our study area (Figure 3a), the gravitational profiles for
 592 this area (Figure 2b-d) shows a pattern similar to that for the area west
 593 of Aceh, but with a substantially smaller near-trench high. The regional
 594 tectonics in this area are further complicated by the transition to active N-S
 595 seafloor spreading behind the accretionary prism in the Andaman Sea. As a
 596 result, whilst the seismicity and gravity profiles are not representative of a
 597 typical subduction zone, without more data we are hesitant to ascribe this
 598 to collapse of the forearc, as we suggest is occurring west of Aceh.

599 **6 Comparison to other subduction systems**

600 Large-scale variations in the incoming sediment thickness to subduction sys-
 601 tems also occur elsewhere on the planet, but the observed pattern of in-
 602 traplate seismicity along the Sumatra margin remains unique. We ascribe
 603 this apparent contradiction to the relatively small proportion of the global
 604 subduction system to have sufficient outer rise seismicity to allow the type
 605 of detailed analysis presented here. Sections of several other subduction
 606 zones around the world, most notably Cascadia and the Chilean margin near
 607 Concepción, show similar forearc morphology variations to that seen west of
 608 Aceh, and have also been suggested to be undergoing forearc collapse (Mc-
 609 Neill et al., 1997; Goldfinger et al., 2000; Geersen et al., 2011). However,
 610 relatively little intraplate seismicity has been observed along these margins
 611 during the instrumental period, and as such the intraplate strain is hard to
 612 assess. Hence, we suggest that when such seismicity does occur, likely in the
 613 period following a major earthquake on the adjacent subduction interface,
 614 the seismicity within the downgoing plate may show a pattern similar to that
 615 that we have observed west of Aceh.

616 Offshore northern Oregon and Washington, margin-perpendicular forearc
 617 extension from the late Miocene to present has produced normal faults within
 618 the sedimentary prism (McNeill et al., 1997). This process is limited to a
 619 region where the incoming plate surface is dominated by the major Astoria
 620 and Nitinat submarine fans, with incoming sediment thicknesses of 3–4 km,

621 tapering away to both the south and north of the collapsing section of the
622 margin (Goldfinger et al., 2012), suggesting that, as we infer for Sumatra,
623 short-timescale variations in incoming sediment thickness can lead to rapid
624 periods of forearc readjustment and collapse. Increased sediment thickness
625 also has the effect of smoothing or masking the structure of the downgoing
626 plate along the plate interface. This has been speculated to be a contributing
627 factor in sustaining large, smooth ruptures during megathrust earthquakes,
628 even in cases where the stress is lowered (Ruff, 1989) – a hypothesis that
629 would fit with the spatial correlation of our suggested region of margin col-
630 lapse with the region of highest slip in the 2004 Aceh-Andaman earthquake.

631 7 Conclusions

632 The seismicity of the near-trench region of the Sunda Arc west of Sumatra
633 shows a notable departure from the global trend, with shallow compressional
634 earthquakes occurring within the downgoing oceanic plate, in a region typ-
635 ically expected to be in horizontal extension. This region coincides with an
636 area in which the forearc prism shows a steep front and low-angle top, char-
637 acteristic of a region undergoing morphological readjustment in response to
638 a change in the boundary conditions governing the shape of the accretionary
639 prism. This change in prism morphology, with the prism propagating out-
640 wards over the downgoing plate, leads to closely-spaced regions of bending
641 and unbending in the downgoing plate. The phase of prism collapse likely
642 results from a rapid change in incoming sediment thickness and viscosity.

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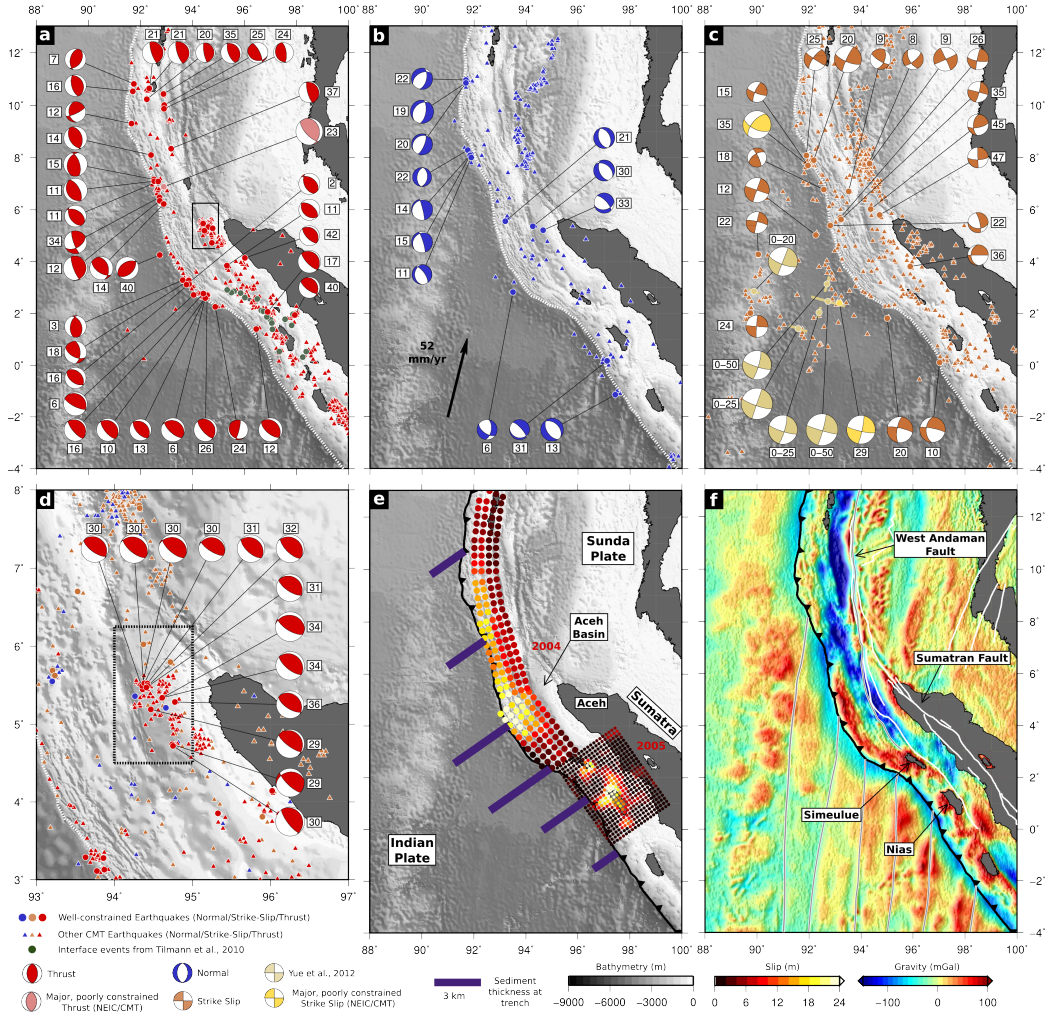


Figure 1: Seismic activity and plate structure west of Sumatra. Earthquakes with well-constrained source parameters from this study are plotted as circles, with associated focal mechanisms. The depth beneath the seabed of each earthquake is given by the number next to the mechanism. Events from the gCMT catalogue are shown as triangles, for those earthquakes occurring within 100 km seawards, and 300 km landwards, of the trench. (a) Thrust-faulting earthquakes. Green points are the low-angle interface events of Tilmann et al. (2010). Earthquakes within the dashed box are shown in (d). (b) Normal-faulting earthquakes. The black arrow is the convergence vector between the Indian plate and the Sunda plate (DeMets et al., 2010). (c) Strike-slip faulting earthquakes. Beige mechanisms are sub-events of the 2012 Indian Ocean earthquake (Yue et al., 2012), with bars indicative of along strike extent of rupture, and depth ranges indicative of the depth range of major slip in finite-fault models. (d) Thrust-faulting earthquakes in the Aceh Basin. (e) Slip models for the 2004 Aceh-Andaman (Rhie et al., 2007) and 2005 Nias (Konca et al., 2007) earthquakes. The slip magnitudes for the Nias event have been multiplied by a factor of 3 relative to the Aceh-Andaman event, to make the two events visible on the same colour scale. Sediment thicknesses seaward of the trench are shown by the thick purple bars (McNeill and Henstock, 2014, and references therein). (f) Free-air gravity anomalies (Sandwell and Smith, 2009). Grey and white lines mark fracture zones in the Indian plate, and major strike slip fault systems in the overriding plate.

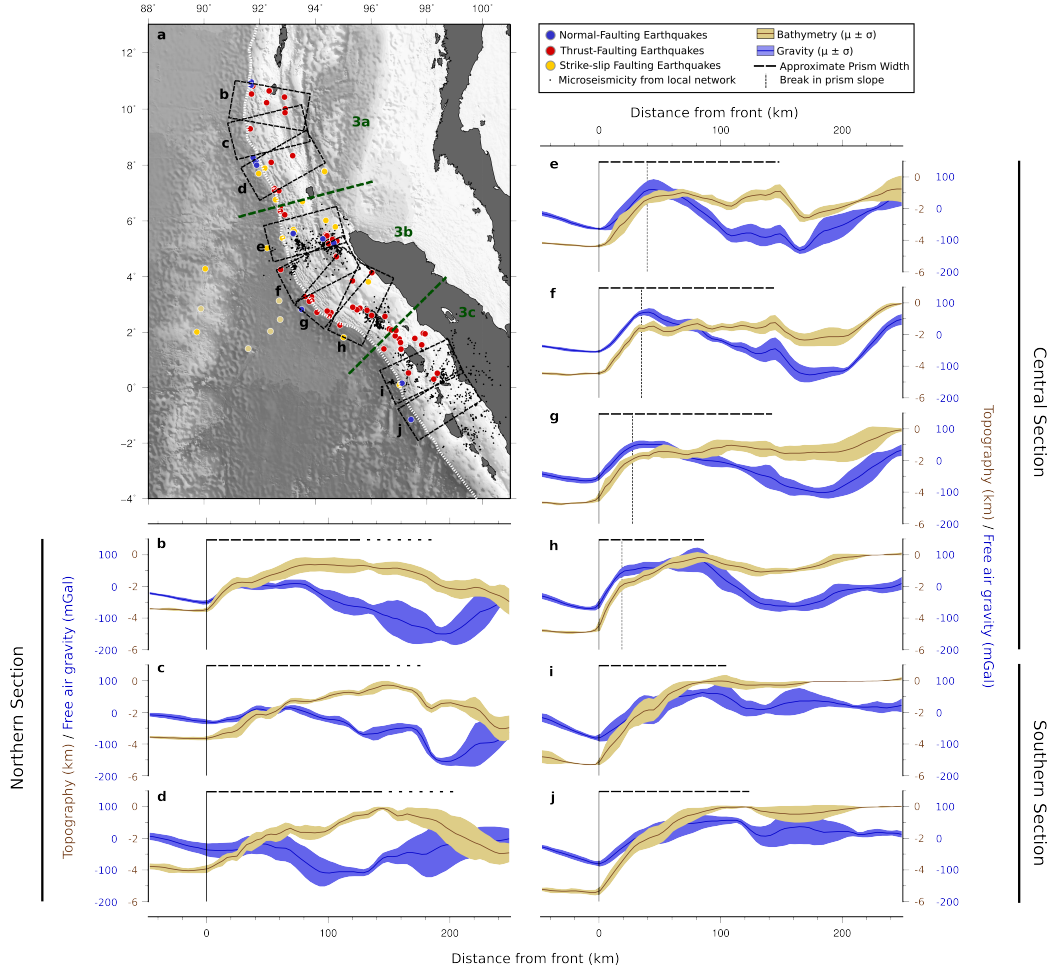


Figure 2: (a) Map of earthquakes with well-constrained depths, coloured by mechanism. Black points are microseismic activity from local seismic deployments (Lin et al., 2009; Tilmann et al., 2010; Lange et al., 2010). Black dashed boxes are the areas used for swaths of bathymetric and gravity data shown in (b) – (j). Green dashed lines separate the regions used for the cross-sections shown in Figure 3. (b) – (j) show mean (darker line) and $\pm 1\sigma$ values (shaded bands) for the trench-perpendicular swaths shown on (a). Beige/brown are for bathymetric/topographic data, blues are for free-air gravity data. Vertical solid lines indicate the location of the trench. Vertical dashed lines on (e) – (j) indicate the principal break in slope. Horizontal dashed lines indicate the approximate prism width in each case (the western prism boundary on (b) – (d) is uncertain).

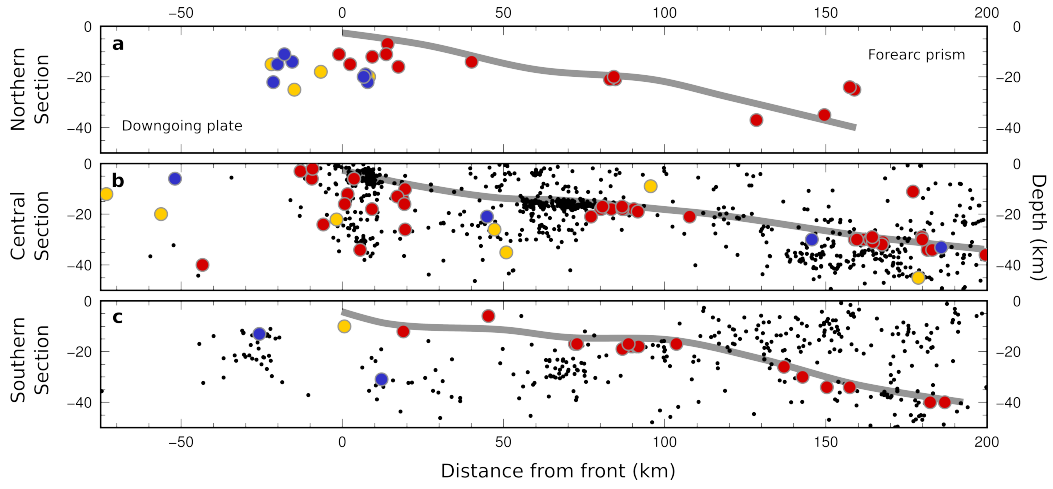


Figure 3: Cross-sections through earthquakes north of the green line on Figure 2a intersecting the trench at 6.5°N (e), between the two green lines (f) and south of the green line intersecting the trench at 1.5°N . All earthquakes are shown at their minimum trench-perpendicular distance. Red points are thrust-faulting earthquakes, blue are normal-faulting earthquakes, and yellow are strike-slip faulting earthquakes (as on Figure 2a). Small black points are earthquake hypocentres from local seismic network deployments, as shown on Figure 2a. Depth is indicative of their depth below sea level.

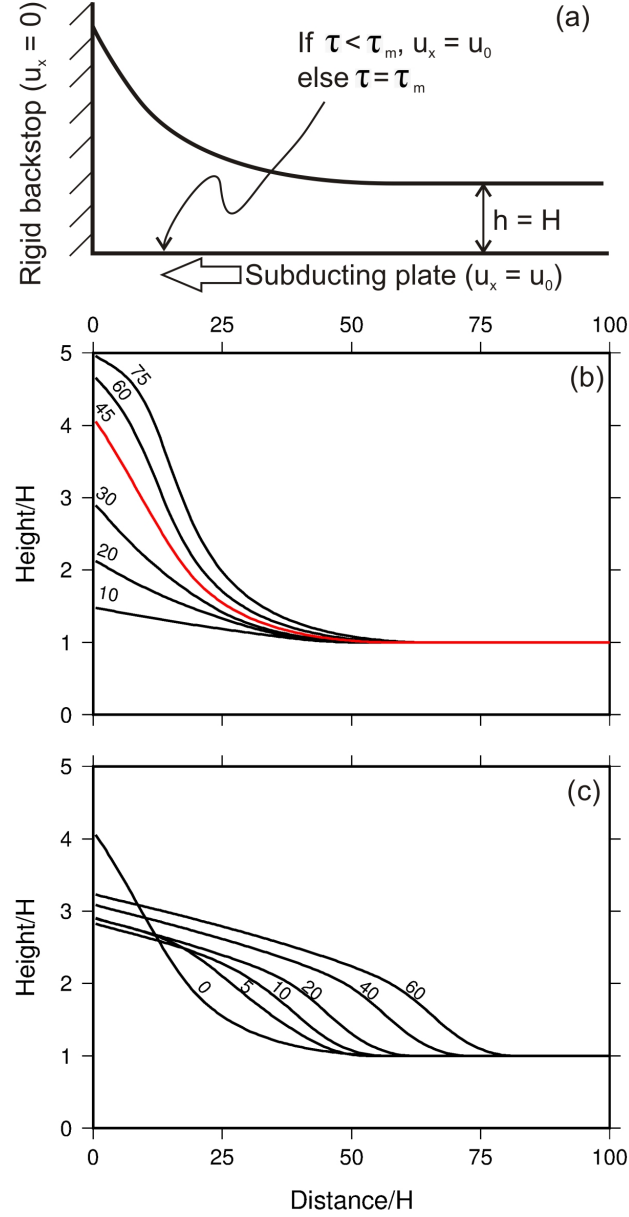
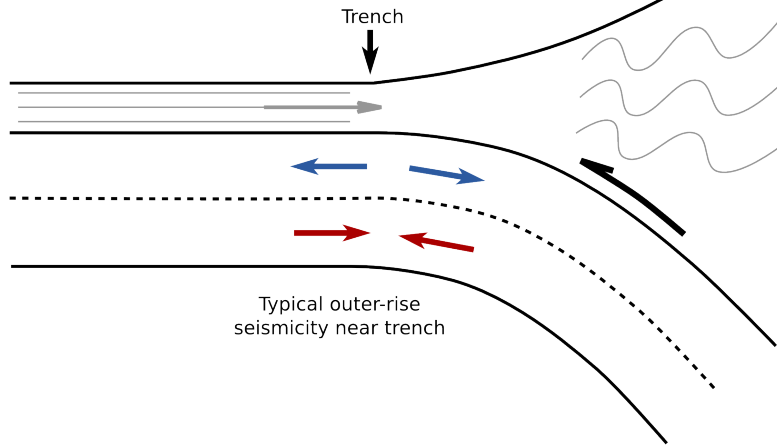


Figure 4: Modelling forearc evolution. (a) model setup. (b) model results for $\alpha=15$ and $\tau_m = 0.05$. These values are equivalent to a wedge viscosity of $2 \times 10^{20} \text{ Pa s}$ and a megathrust shear stress of 16 MPa, for a convergence rate of 52 mm/yr, sediment thickness of 1 km, and density of 2500 kg/m^3 . This viscosity is similar to that which Copley and McKenzie (2007) found for the onshore Indo-Burman sedimentary wedge, and the shear stress is similar to the stress-drops observed in megathrust earthquakes. The curves show the topography labelled with non-dimensionalised time. For the parameters chosen, a non-dimensional time of 45 is equivalent to $\sim 900 \text{ kyr}$. (c) Model results when the red curve in (b) is taken as a starting configuration, and the value of α is reduced by a factor of 10.

a) Equilibrium forearc



b) West of Aceh

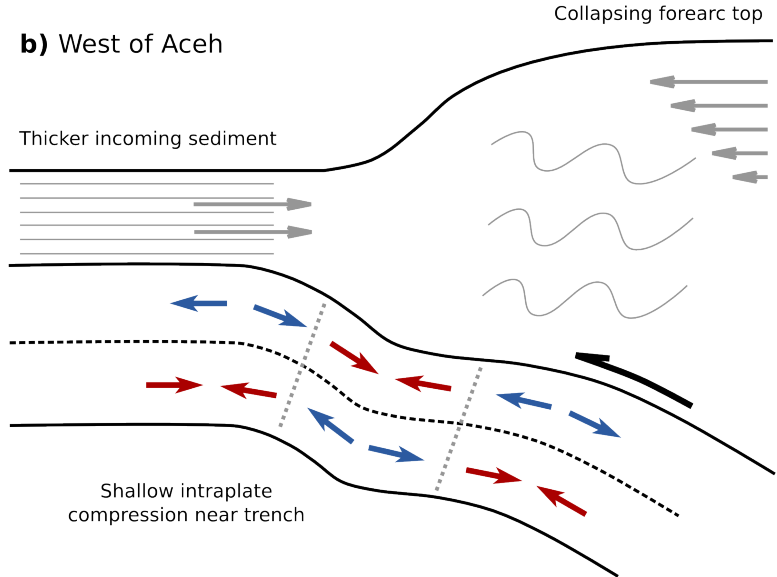


Figure 5: Schematic diagram linking forearc morphology and bending strains within the downgoing plate. (a) The globally-typical scenario for forearcs in equilibrium. (b) The scenario we propose for the Sunda arc west of Aceh during forearc readjustment.